INTRODUCTION

This report summarizes analysis of pumping test data from a production well completed in the deep volcanic aquifer at the Big Sandy Energy Project site near Wikieup, Arizona. The pumping test was conducted to characterize the hydraulic properties of the aquifer and to determine its suitability for supplying water for a gas fired power plant over the 40-year anticipated life of the project. It is expected that the plant will require a water supply of 3000 gallons per minute (gpm) for that period.

Exploration of the area has identified three separate aquifers at the site: an Upper Aquifer consisting of the Upper Basin fill and Recent Stream and Flood Plain deposits; a Middle Aquifer consisting of the Lower Basin fill; and a deep volcanic aquifer referred to as the Lower Aquifer.

Tight lakebed clays up to hundreds of feet or more in thickness separate the Upper and Middle Aquifers, ensuring their hydraulic separation throughout most of the area. In the vicinity of the pumping test, a thin volcanic layer (10 feet thick) separates the Lower (volcanic) Aquifer from the Middle Aquifer. This thin layer apparently provides hydraulic separation of the Middle and Lower Aquifers in the test area. Some of the Lower Aquifer wells flow and have shut in pressures of nearly 100 feet of head, while the water levels in the Middle Aquifer wells are below land surface.

Geologic mapping and hydraulic evidence suggest that the volcanic aquifer has limited areal extent, currently estimated to be about 57 square miles. The aquifer dips to the west and rises to the east toward its recharge area. The higher elevation of the recharge area accounts for the observed artesian pressure in the aquifer.

PW2 was installed as the pumped well for the aquifer test. It is a 20-inch cased well with 12-inch screen set from 1135 feet to 1600 feet below land surface. Lower Aquifer observation wells monitored during the test included OW2, OW3 and OW4, located 200 feet, 4880 feet and 3150 feet, respectively, from the pumped well. Other observations wells used during the test included:

1. Middle Aquifer Well OWMA2, located 200 feet from the pumped well
2. Upper Aquitard Well OW8, located nearly a mile from the pumped well

3. Upper Aquifer Well OW7, about a half mile away
4. Upper Aquifer Banegas Well, over one mile away
5. Upper Aquifer Harris Well, over one and a half miles away
6. Upper Aquifer Well OW1, over one and a half miles away

Test pumping of PW2 began at 3:30 PM on September 11, 2000 and continued for 15,720 minutes (10.916 days). Following shutdown, water levels were monitored for an additional 14,139 minutes (9.819 days). The pumping rate initially was about 2000 gpm, but declined to around 1950 gpm by the end of the test. The average pumping rate throughout the test was 1960 gpm.

For a complete description of the site layout, geology, well construction and pumping test details, the reader is referred to the report entitled *Water Resources of the Southern Portion of the Big Sandy Valley, Wikieup, Mohave County, Arizona* by Manera, Inc.

TIME-DRAWDOWN AND RECOVERY DATA

Time-drawdown and recovery data from the pumped well and Lower Aquifer wells OW2, OW3 and OW4 were plotted on various graphs for analysis. This section presents several of the data plots and offers an interpretation of the pumping data response.

It should be pointed out that the data from this test reflected a very different response than observed in most aquifer pumping tests. It is believed that there were two causes of the unusual hydraulic response. First, the lateral extent of the aquifer is fairly limited and, thus, boundary effects tended to dominate the pumping and recovery data. Second, the aquifer likely does not conform to "porous media" aquifer assumptions.

In addition to weathered unconsolidated inter-bedded zones, the makeup of volcanic aquifers often includes major voids, passageways, joints, fractures and conduits along which preferential groundwater flow occurs. These openings can have enormous hydraulic conductivity and spread the influence of pumping very quickly. After initial drawdown response in the high permeability openings, water begins to move from the less permeable portions of the aquifer into the fractures. Thus the response to pumping is analogous to the "block and fracture" response hypothesized for fractured aquifers.

Both porous media and fractured rock analytical approaches were used to evaluate the test data. As described below, the pumping test results were consistent with a bounded aquifer having highly transmissive fractures connected to moderately transmissive blocks.

Figures 1 through 4 show semilog time-drawdown graphs for the pumped well PW2 and observation wells OW2, OW3 and OW4, respectively. Figures 5 through 9 show corresponding recovery plots for these wells. Figures 5 and 6 both show recovery data from the pumped well; Figure 5 shows the complete data set while Figure 6 shows an expanded scale to make the slope of the data plot more readily discernable.

Aquifer parameters were calculated following the conventional porous medium approach using the Cooper-Jacob equation. According to this method, transmissivity is calculated from the time-drawdown and recovery graphs using:

$$T = \frac{264Q}{\Delta s}$$

and storage coefficient is calculated from the time-drawdown graphs as follows:

$$S = \frac{0.3Tt_0}{r^2}$$

where,

T = transmissivity, in gallons per day per foot (gpd/ft)
S = storage coefficient
Q = discharge rate, in gpm
Δs = change in head over one log cycle of the graph, in feet
t_0 = the zero drawdown intercept of the straight line of best fit, in days
r =distance from the pumped well, in feet

The calculated parameters for each well are shown on the respective graphs. Inspection of the values suggests that the calculations provided erroneous results. The primary evidence of this is the set of values of calculated storage coefficient. They were not consistent from well to well, but decreased with increasing distance from the pumped well. The very large value (0.33) obtained from OW2 strongly suggests the presence of a negative boundary.

Note also that the time-drawdown graphs for OW2, OW3 and OW4 were nearly identical, even though the range of distances from the pumped well spanned more than an order of magnitude. This response was consistent with all three wells penetrating the same high permeability fracture or joint system.

The data suggest that the cone of depression expanded very rapidly during the test. Drawdown was observed in OW3, nearly a mile from the pumped well, within the first few minutes of pumping. Also, a flow reduction in Cofer Hot Springs, two and a half miles from the production well, was observed in response to pumping. Thus, it is possible that the cone of depression in the fractures could have reached the boundary limits of the volcanic aquifer within the first several hours of pumping. This would account for the steep slopes on the time-drawdown graphs, which subtly continued to increase throughout the test.

A comparison of Figures 1 and 6 for the pumped well shows a steeper slope on the time-drawdown graph than on the recovery graph. The pumped well time-drawdown slope also was steeper than the slopes on the observation well time-drawdown graphs. Theoretically, all of these slopes should be identical. The steeper pumped well time-drawdown slope indicates a phenomenon seen occasionally in pumping tests and is usually caused by a gradual reduction in hydraulic conductivity around the well bore (well efficiency reduction) during pumping.

DISTANCE-DRAWDOWN DATA

Figure 10 shows a distance-drawdown-like plot of the observation well data. On this graph, distance has been plotted versus calculated recovery. Calculated recovery is determined as the difference between the actual remaining drawdown observed after pumping stopped and the drawdown, extrapolated from the time-drawdown graph, that would have resulted had pumping continued. Aquifer parameters from this graph were calculated as follows:

$$T = \frac{528Q}{\Delta s}$$

$$S = \frac{0.3Tt}{R^2}$$

where,

T	= transmissivity, in gallons per day per foot (gpd/ft)
S	= storage coefficient
Q	= discharge rate, in gpm
Δs	= change in head over one log cycle of the graph, in feet
t	= recovery time, in days
R	=log-extrapolated radius of influence, in feet

Note that the calculated transmissivity exceeded three million gallons per day per foot (gpd/ft) and the storage coefficient was about 10^{-43}, an impossible value. The low storage coefficient confirms the negative boundary while the high calculated transmissivity value suggests very highly permeable conduits in the volcanic aquifer. Keep in mind that the calculated value may not reflect the true transmissivity, because it was obtained by applying porous media methods to an aquifer that exhibits fractured rock response.

EARLY-TIME RESPONSE

An attempt was made to analyze early recovery data to try to avoid the effect of the negative boundary on the response data. To accomplish this, log-log plots were made of recovery versus time for OW2, OW3 and OW4, as shown on Figures 11, 12 and 13, respectively. Recovery data were used because the early data are immune to the effects of pumping rate variations, unlike early pumping data.

Theis curve matching was applied to the early data to calculate aquifer parameters. Theis curve matching is performed using the Theis type curve – a plot of the Theis well function *W(u)* versus *1/u*. Curve matching is accomplished by overlaying the type curve on the data plot and, while keeping the coordinate axes of the two plots parallel, shifting the data plot to align with the type curve, effecting a match position. An arbitrary point, referred to as the match point, is selected from the overlapping parts of the plots. Match point coordinates are recorded from the two graphs, yielding four values –

W(u), *1/u*, *s* and *t*. Using these match point values, transmissivity and storage coefficient are computed as follows:

$$T = \frac{114.6Q}{s} W(u)$$

$$S = \frac{Tut}{2693r^2}$$

where,

T	= transmissivity, in gpd/ft
S	= storage coefficient
Q	= discharge rate, in gpm
$W(u)$	= match point value
s	= match point value
u	= match point value
t	= match point value
r	= distance from pumped well, in feet

The results showed an average transmissivity of nearly three million gpd/ft, similar to the value obtained from the distance-recovery analysis. The average storage coefficient calculated from these graphs was just over 10^{-3}, a reasonable value.

LATE-TIME RESPONSE

Figures 14, 15 and 16 show linear plots of the time-drawdown data for OW2, OW3 and OW4, respectively. Quite significantly, these graphs showed a "straight line" or linear time-drawdown relationship over the last five days of pumping. The only way such a response is possible, whether applying porous media theory or fractured rock theory, is for the cone of depression to be fully expanded to the limits of the aquifer boundaries. An analogy is the water level response that would be observed when pumping a bathtub dry. These data plots confirm the rapid expansion of the cone of depression to the aquifer boundaries and the limited lateral extent of the aquifer.

AQUIFER EXTENT

The linear rate of descent of the cone of depression was used along with storage coefficient estimates to approximate the lateral extent of the volcanic aquifer. This was done to see if observed data were consistent with the previously estimated aquifer area of 57 square miles.

The storage coefficient of an aquifer includes two components: S_w, associated with expansion (decompression) of the water in the aquifer during pumping, and S_f, associated with elastic contraction of the aquifer skeleton in response to water pressure reduction caused by pumping. For groundwater at 96 degrees Fahrenheit, 2.32 feet of water exerts a pressure of 1 psi and, thus:

$$S_w = \frac{nb}{2.32C_w}$$

where,

S_w = storage coefficient component caused by water expansion
n = aquifer porosity (assumed to be 0.10)
b = aquifer thickness (assumed to be 500 feet)
C_w = compressibility of water (323,000 psi at prevailing temperatures)

Using these inputs, S_w equals 6.7×10^{-5}. Generally, S_f is substantially greater than S_w, typically one half to one order of magnitude or more. Thus, a reasonable range for the aquifer storage coefficient might be expected to be approximately 4×10^{-4} to 1.2×10^{-3}.

An expression relating aquifer area to storage coefficient was developed using the observed rate of decline of water levels during late pumping and the observed rate of water level rise during late recovery. The late-time rate of decline was determined to be 0.318 feet per day from Figures 14 through 16. Figures 17, 18 and 19 show linear plots of recovery data for OW2, OW3 and OW4, respectively. The rate of recovery at the end of the measurement period was estimated from the graphs to be 0.103 feet per day.

In developing this analysis, a simplifying assumption was made of a constant recharge rate to the aquifer during pumping and recovery. Prior to the pumping test, there was no *net* recharge to the aquifer. The volume of water recharging the aquifer was presumably offset by the volume of water discharging to springs, seeps and adjacent aquifers. During pumping and recovery, the lowering of the water levels in the aquifer would be expected to reduce groundwater outflow (and possibly increase groundwater inflow). This net recharge caused by the pumping test was assumed to occur at a constant rate, Q_{in}. This is clearly an over-simplification and, therefore, makes the following analysis only approximate. The analysis is, nevertheless, worthwhile to provide a reality check on the estimated aquifer area.

During late pumping time, applying the definition of storage coefficient gives the following:

$$S = \frac{Q - Q_{in}}{Ad}$$

and during late recovery time:

$$S = \frac{Q_{in}}{Ar}$$

where,

S	= storage coefficient
Q	= pumping test discharge rate
Q_{in}	= net flux into aquifer caused by pumping
A	= area of aquifer
d	= rate of descent of water level during late pumping time
r	= rate of rise of water level during late recovery time

Solving these two equations for A yields:

$$A = \frac{Q}{S(d + r)}$$

Expressing A in square miles and Q in gpm:

$$A = \frac{192.5Q}{5280^2 S(d+r)}$$

Using this last equation, a graph was made of A as a function of S and is shown on Figure 20. As indicated, reasonable estimates for storage coefficient placed the estimated aquifer area between approximately 25 and 80 square miles, in good agreement with initial estimates. Even though the foregoing analysis can only be expected to yield approximate results, it provides additional confidence that previous estimates of aquifer size are reasonable.

LINEAR FRACTURE ANALYSIS

A simplified fracture analysis method was applied to the drawdown data to provide additional information on aquifer properties. According to Jenkins and Prentice (*Ground Water*, January-February 1982), the drawdown in observation wells penetrating the same pumped fracture can be approximated by:

$$s = \frac{Q}{L\sqrt{\pi TS}}\sqrt{t}$$

where,

Q = discharge rate, in cubic feet per day
L = fracture length, in feet
T = transmissivity, in square feet per day
S = storage coefficient
t = pumping time, in days

This relationship implies that a plot of drawdown versus the square root of time will be a straight line with slope, m, as follows:

$$m = \frac{Q}{L\sqrt{\pi TS}}$$

Solving for T:

$$T = \frac{Q^2}{\pi S m^2 L^2}$$

Time-drawdown data from OW2, OW3 and OW4 were plotted versus the square root of time and are shown on Figures 21, 22 and 23, respectively. The average slope from these graphs was determined to be 8.17 feet per 130 minutes$^{1/2}$, or 2.38 feet per day$^{1/2}$. Converting the units of the discharge rate of 1960 gpm yields 377,300 cubic feet per day. To determine a fracture length to use in the above equation, it was assumed that the fracture spanned the entire aquifer domain. Thus, a fracture length of 39,860 feet was used in the equation. This is the length dimension of a square having an area of 57 square miles, the presumed size of the aquifer. Finally, from Figure 20, the storage coefficient corresponding to an aquifer area of 57 square miles is 5.6 x 10^{-4}.

Inputting these values into the expression for T yielded a value of 8990 square feet per day, or 67,300 gpd/ft. This value can be thought of as approximating the transmissivity of the "blocks" of aquifer material, which, in turn, are connected to more highly transmissive fractures. The block transmissivity value may help explain the relatively modest specific capacity of the pumped well. PW2 was pumped at 1960 gpm with 151 feet of drawdown making the specific capacity only 13 gpm per foot of drawdown. This value is very low compared to the fracture transmissivity values obtained from the pumping test analysis, but is in line with expected well yields based on the block transmissivity value. (Note that the block transmissivity varies as the inverse square of the fracture length and, thus, a wide range of values could be obtained for different assumptions of fracture length.)

INTERCONNECTION OF AQUIFERS

Part of the objective of the pumping test was to evaluate the inter-connection between the Lower Aquifer and the Middle and Upper Aquifers. To that end, water levels were monitored in several Middle Aquifer, Upper Aquitard and

Upper Aquifer wells as mentioned earlier. Only pumping data were available from most of the wells. Plots of time-drawdown data from these wells are available in the Manera, Inc. report. A brief summary of the response of each of the monitored wells follows.

<u>Middle Aquifer well OWMA2 (200 feet from PW2)</u>
Water levels in OWMA2 rose steadily throughout the pumping period about half a foot. This could be either a continuation of a pre-pumping water level trend or an indication of poroelastic response to pumping (the so-called Noordbergum effect). The lack of a reversal of the upward water level trend shows hydraulic separation of the middle and lower aquifers at this location.

<u>Upper Aquitard well OW8 (less than one mile from PW2)</u>
Water levels in OW8 trended upward slightly with a couple of unexplained upward steps or discontinuities. The data showed no hydraulic connection between the Upper Aquitard and the Lower Aquifer.

<u>Upper Aquifer well OW7 (one half mile from PW2)</u>
Water levels in OW7 showed no significant response to pumping.

<u>Upper Aquifer Banegas well (over one mile from PW2)</u>
Water levels in the Banegas well showed no response to pumping PW2. During the pumping period, water levels rose about an inch and then declined about an inch and a half, continuing a sinusoidal pattern observed prior to the pumping test.

<u>Upper Aquifer well OW1 (over one and one half miles from PW2)</u>
Water levels in OW1 declined about one half inch at a steady rate during the pumping test.

<u>Upper Aquifer Harris well (over one and one half miles from PW2)</u>
Water levels in the Harris well declined approximately two inches at a steady rate during the pumping test.

As a group, these wells did not show a consistent response to pumping Lower Aquifer well PW2. Keep in mind that it is not necessary to pull the Lower Aquifer water levels below levels in the overlying aquifers to induce a hydraulic response. If a sufficient hydraulic connection existed between the Lower Aquifer and Middle Aquifer, for example, pumping the Lower Aquifer could induce drawdown in the Middle Aquifer, even if water levels in the Lower Aquifer were higher than those in the Middle Aquifer. This is because pumping the Lower Aquifer would disturb the equilibrium between the aquifers, thereby reducing the pre-existing rate of groundwater flux from the Lower Aquifer to the Middle Aquifer.

The lack of a discernable hydraulic response between the Middle and Lower Aquifers places an upper limit on the possible vertical hydraulic conductivity of the aquitard separating them. Obtaining this limiting value analytically is extremely difficult. It is best determined via numerical modeling. For example, the vertical hydraulic conductivity of the aquitard can be increased systematically during model calibration until a "measurable" response in the unpumped aquifer is predicted by the model. The actual vertical hydraulic conductivity of the aquitard then must be below this threshold value.

The only possible exceptions to the lack of response to pumping PW2 were the water level trends observed in OW1 and the Harris well. Water levels in both of these wells declined slightly in a uniform fashion during pumping, not unlike what would be expected if there were a hydraulic connection between the Upper and Lower aquifers. To evaluate this response, a comparison was made of the pumping response in these wells and background data recorded prior to and after pumping.

Figure 24 shows a hydrograph for the Harris Well starting on May 30 and continuing until October 2, 10 days after shut down of PW2. The hydrograph showed a clear downward water level trend throughout the entire monitoring period. Note that data recorded during pumping (September 11 through 22) and recovery (September 22 through October 2) showed no deviation from the overall background trend. Thus, the gradual water level decline observed during the pumping test simply reflected the background water level decline already in progress. These data demonstrate hydraulic separation of the Lower and Upper aquifers in the vicinity of the Harris Well.

Figures 25 and 26 show manually recorded water levels in Upper Aquifer wells OW1 and the Salazar Well, located just 400 feet from OW1. The Salazar Well was monitored during the month of August whereas OW1 was monitored during the 21 days of pumping and recovery of PW2, from September 11 until October 2. The manual measurements were recorded to the nearest tenth of a foot, resulting in rather uneven data plots. Nevertheless, it was possible to discern a gradual decline in water levels in this area of the Upper Aquifer for the entire monitoring period from August to October. Thus, the observed decline in water levels in OW1 during the pumping test was likely a result of the overall background trend rather than a response to pumping PW2. Also, inspection of Figure 25 did not show a clear water level decline followed by a water level rise in response to pumping and recovery of PW2. Thus, the data from OW1 and the Salazar Well support the thesis of hydraulic separation of the Upper and Lower aquifers in this area of the site.

It also has been hypothesized that the Lower Aquifer might be in hydraulic communication with an underlying unit termed the Arkosic Gravel, a highly transmissive and laterally extensive formation. However, the pumping test data were not consistent with this idea.

Inspection of the time-drawdown graphs on Figures 2, 3 and 4 shows that the slopes of these plots increased continuously with time. Expansion of the cone of depression into a laterally extensive aquifer would have resulted in stabilized slopes and, thus, the ever-increasing slopes suggest a laterally limited aquifer.

Similarly, the *linear* drawdown plots on Figures 14, 15 and 16 showed stabilized slopes, consistent with a laterally limited aquifer. Expansion of the cone of depression into a laterally extensive aquifer would have resulted in a gradual flattening of these slopes.

Finally, the recovery data plotted on Figures 6, 7, 8 and 9 show an extrapolated trend that would leave the recovered levels below the original static water level. This is often an indication that the cone of depression has reached the aquifer boundaries during pumping. Thus, the pumping and recovery data were consistent with pumping from an areally limited aquifer. This implies that the Arkosic Gravel is either limited in areal extent or hydraulically separated from the Lower Aquifer.

NO POROSITY DATA

It is important to point out that the pumping test does not provide (and cannot provide) any information on the porosity of the aquifer. The two key aquifer characteristics that determine the volume of water stored in the volcanic aquifer are areal extent and porosity. While the pumping test data were able to support confirmation of the approximate size of the aquifer, there is unfortunately no way to obtain porosity information from the test results.

CONCLUSIONS

The following conclusions can be drawn from the pumping test data analysis.

1. The aquifer response to pumping exhibited characteristics of both porous media response and fractured aquifer response.
2. Hydraulic response was consistent with either a highly transmissive porous media aquifer or a fractured aquifer with highly transmissive fractures and moderately transmissive blocks.
3. Most of the pumping response reflected the effects of aquifer boundaries.
4. A reasonable estimated range of storage coefficient is about 4×10^{-4} to 1.2×10^{-3}.
5. The hydraulic response to pumping was consistent with an aquifer having an area of approximately 25 to 80 square miles.
6. Hydraulic response was equivalent to what would be expected from a laterally limited porous media aquifer having a transmissivity in excess of 10^6 gpd/ft and a storage coefficient around 10^{-3}. It is also consistent with a fractured rock aquifer having fracture transmissivity over 10^6 gpd/ft and block transmissivity of 67,000 gpd/ft.
7. Linear drawdown response during the last several days of pumping indicated that the cone of depression was fully developed throughout the entire lateral extent of the aquifer.
8. The efficiency of the pumped well appeared to decline somewhat during pumping.
9. The data suggest that the Lower Aquifer is hydraulically separated from the Middle Aquifer and Upper Aquifer.
10. The data suggest that if the Arkosic Gravel is laterally extensive, it is not in hydraulic communication with the Lower Aquifer